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# Present-day lithospheric strength of the Eastern Alps and its relationship to neotectonics

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[1] We calculate the present-day lithospheric strength of the Eastern Alps along the new reflection seismic profile TRANSALP to examine vertical and lateral strength variations and their implications on neotectonic activity of the Eastern Alps. The large-scale geometry of the Eastern Alps and the spatial distribution of upper, and lower crustal layers, and the lithospheric mantle is constrained by the deep seismic line. Two rheological models, coupled to a kinematic thermal model that accounts for the thermal evolution of the Eastern Alps for the last 30 Myr, are investigated for the present-day lithospheric configuration in the Eastern Alps. Models with strong (Model A) and weak (Model B) crustal rheologies predict the European and the Adriatic plates to be stronger than the central zone of the orogen comprising the region between the Inntal Fault and the Periadriatic Fault. Model A is characterized by a brittle-ductile boundary between 14 and 9 km depth and strong coupling of the mechanically strong lower crust to the upper mantle, whereas Model B suggests the presence of a thick decoupling zone between the upper crust and the upper mantle and a shallower brittle-ductile boundary (7–10 km). Of these end-member scenarios, Model A is in better agreement with neotectonic data including seismicity down to the upper-lower crust boundary within the Adriatic plate, uplift of the central zone of the Eastern Alps and the Southern Alps, and eastward escape of fault-bound blocks. Such deformation pattern is best explained by lateral extrusion upon north-south compression supporting a strong-weak-strong configuration of tectonic units along the TRANSALP line. *INDEX*

*TERMS:* 9335 Information Related to Geographic Region: Europe; 8102 Tectonophysics: Continental contractional orogenic belts; 8107 Tectonophysics: Continental neotectonics; 8159 Tectonophysics: Rheology—crust and lithosphere; 8164 Tectonophysics: Stresses—crust and lithosphere; *KEYWORDS:* Eastern Alps, lithospheric strength, neotectonics, rheology, TRANSALP. *Citation:* Willingshofer, E., and S. Cloetingh, Present-day lithospheric strength of the Eastern Alps and its relationship to neotectonics, *Tectonics*, 22(6), 1075, doi:10.1029/2002TC001463, 2003.

## 1. Introduction

[2] Rocks exposed in the Eastern Alps of central Europe retain the memory of their Paleozoic to present-day evolution. The current geometry of the eastern Alps, however, is largely controlled by the combined affect of surface and subsurface processes since the onset of continent-continent collision during the Oligocene, ca. 35 Ma ago. Collisional coupling between the down-going European plate and the overriding Adriatic plate induced horizontal shortening within the orogenic wedge in-between. Through time, N-S shortening was accommodated by thrusting and folding in the area of maximum compression and by orogen-parallel extension leading to the lateral translation (extrusion) of fault-bound blocks toward the east [Ratschbacher *et al.*, 1991a; Frisch *et al.*, 1998; Neubauer *et al.*, 2000a; Linzer *et al.*, 2002]. The processes held responsible for the buildup of the Alps have been deduced from the analyses of deformation structures on various scales, which allow drawing conclusions on the relative strength of the tectonic units involved [e.g., Ratschbacher *et al.*, 1991a]. Analogue experiments, designed to study the process of lateral extrusion in the Eastern Alps, emphasized the importance of lateral variations of lithospheric strength in such a system [Ratschbacher *et al.*, 1991b]. These results have been confirmed by thermo-mechanical modeling of the Cenozoic evolution of the Eastern Alps [Genser *et al.*, 1996].

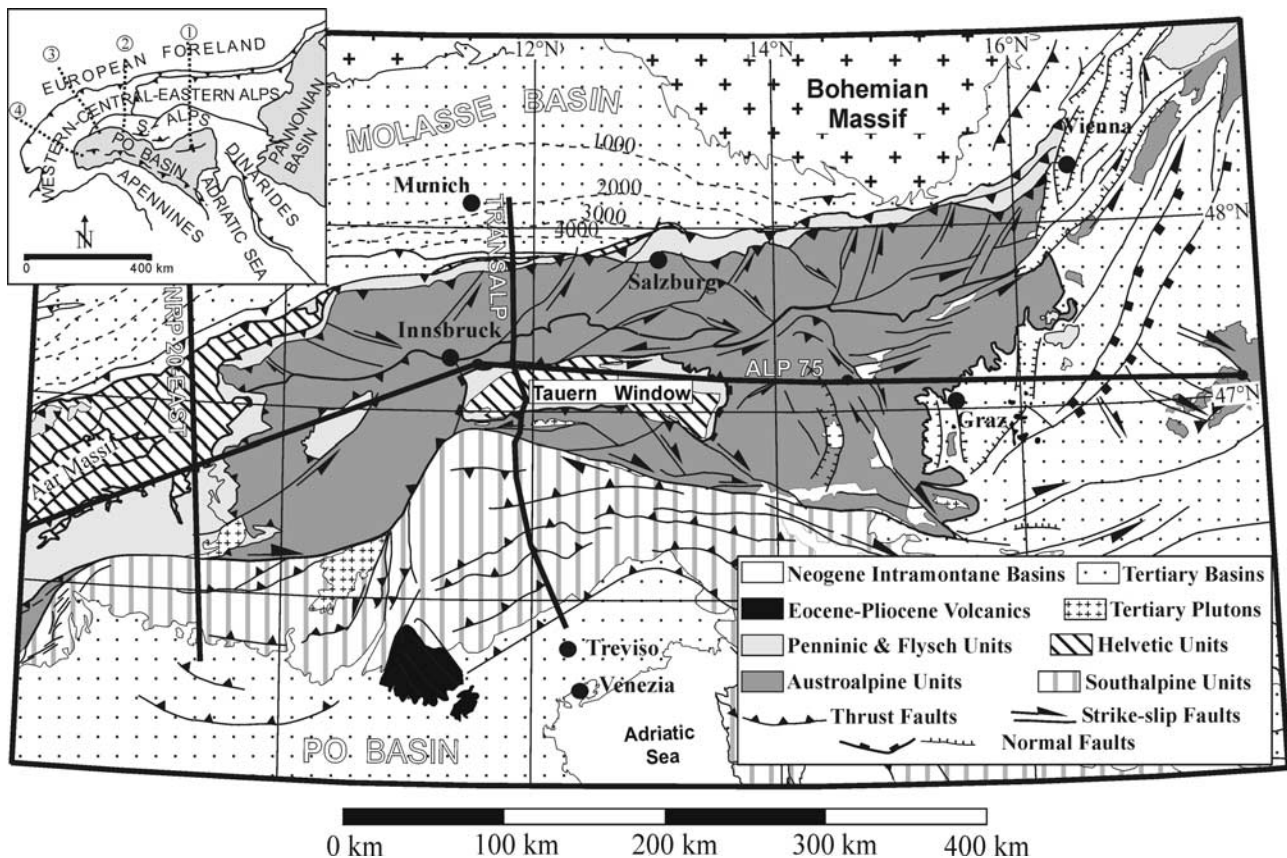
[3] In this paper we present calculations for the present-day lithospheric strength of the Eastern Alps along the new TRANSALP reflection seismic line (Figure 1). The results allow a qualitative and quantitative description of the mechanical state of the lithosphere across the orogen, which is subsequently used as a framework for the discussion of the present-day deformation in the Eastern Alps and its implication on tectonic models.

## 2. Tectonic Setting

[4] Below we give a concise description of the large-scale tectonic units of the Eastern Alps putting emphasis on their thermal evolution since temperature exerts a first order control on the mechanical properties of the lithosphere.

### 2.1. European Plate

[5] The European plate, including the Northern Alpine foreland basin, represents the lower plate in the Alpine Mountain Chain of central Europe with respect to the main body of the Alps, the orogenic wedge, and the Adriatic



**Figure 1.** Tectonic map of the Eastern Alps showing the location of the TRANSALP deep seismic reflection and the ALP75 refraction lines. Paleogeographic units after *Froitzheim et al.* [1996]. Sediment thicknesses of the Molasse Basin are in meters. Inset shows position of reflection seismic profiles across the European Alps. 1-TRANSALP; 2-NRP20-EAST (EGT); 3-NRP20-WEST; 4-ECORS-CROP.

plate [e.g., *Schmid et al.*, 1996; *Neubauer et al.*, 2000b]. Since its last widespread thermal resetting during the Variscan orogeny the European plate was repeatedly affected by spatially limited thermal activity in context of Jurassic and Early Cretaceous formation of the Ligurian-Piemontais (also frequently referred to as “South Penninic Ocean”) and Valais oceans [*Frisch*, 1979; *Froitzheim et al.*, 1996]. These basins are separated in the central and Western Alps by a continental unit the “Briançonnais terrane,” which is missing in the Eastern Alps supporting the idea that the two afore mentioned basins merge somewhere west of the Tauern Window [*Froitzheim et al.*, 1996]. Opening of the Ligurian-Piemontais Ocean resulted in the separation of European derived from Apulian derived (Austroalpine units) paleo-geographic domains [*Channell and Kozur*, 1997]. The locus of extension shifted northward during the Early Cretaceous causing the fragmentation of the European margin leading to the opening of the Valais Basin. In the Eastern Alps remnants of this basin are exposed in the Rhenodanubian Flysch Zone and the Tauern Window, where they overly an intensively deformed fragment of the European continental margin, the “Zentralgneiss unit” (note that the term “Zentralgneiss unit” is used encompassing the polymeta-

morphic basement, the “Zentralgneiss” s. str. (Carboniferous intrusions) and the cover sequences) [*Froitzheim et al.*, 1996]. Subsequent convergence led to the accretion of these units, which are now part of the orogenic wedge. Since the Eocene the European plate was loaded by the orogenic wedge, which progressively moved outward forcing the European plate to bend and underthrust the Alps [*Bachmann and Müller*, 1991; *TRANSALP Working Group*, 2002].

## 2.2. Orogenic Wedge

[6] Here, the “orogenic wedge” comprises the stacked and deformed units between the frontal Alpine thrust in the north and the Periadriatic Fault in the South (Figure 1). From top to bottom the wedge consists of tectonic units of Adriatic (Austroalpine units) and European origin including the stacked cover units of the Helvetic zone and the Zentralgneiss unit in the Tauern Window. These major units are separated by the Flysch Zone and the Penninic suture containing remnants of the Ligurian-Piemontais Ocean. Unlike the Austroalpine units, which experienced mainly Cretaceous metamorphism, rocks exposed in tectonic windows along the central axes of the Eastern Alps, record Tertiary thermal activity [e.g.,

*Frey et al.*, 1999]. In particular, rocks of the suture zone have been metamorphosed under eclogite facies conditions probably during the Late Cretaceous to Paleocene. Subsequent blueschist and amphibolite facies metamorphic overprint was coeval with amphibolite facies metamorphism within the continental Zentralgneiss unit during the late Eocene-Oligocene [*Neubauer et al.*, 2000b, and references therein]. Peak metamorphic conditions are related to the decompression history of the Zentralgneiss unit and slightly predate lateral extrusion tectonics, which climaxed during the Miocene and also caused a fragmentation of the overlying Austroalpine units [*Ratschbacher et al.*, 1991a].

[7] Additionally, a heat source different to metamorphism was provided by the emplacement of plutons close to the Periadriatic Fault and within the orogenic wedge south of the Tauern Window. These fault-controlled plutons intruded the bedrock during the late Eocene to early Oligocene [e.g., *Schmid et al.*, 1996; *Müller et al.*, 2000] and are interpreted as a result of slab breakoff [*Davies and von Blanckenburg*, 1995].

### 2.3. Adriatic Plate

[8] The Adriatic plate represents the upper plate in the Alpine Mountain chain and acted as an interder during the final stages of collision. After the Variscan thermal overprint, which reached greenschist to amphibolite facies grade [*Frey et al.*, 1999] the Adriatic plate was affected by extensional tectonics ultimately leading to the opening of the Ligurian-Piemontais Ocean during the Late Triassic to early Late Jurassic time period [e.g., *Bertotti et al.*, 1993]. The extent of the rifting-related thermal perturbation appears to be more pronounced in western than in eastern parts of the Southern Alps [*Bertotti et al.*, 1999]. The post-Mesozoic evolution of the eastern Adriatic plate reflects repeated phases of contraction leading to the imbrication of the Mesozoic cover sequences, represented by the Dolomite Mountains of the Southern Alps [e.g., *Castellarin and Cantelli*, 2000 and references therein]. Most importantly, Eocene to recent southward directed thrusting on the Adriatic plate was not associated with metamorphism. Contemporaneous with early phases of thrusting in the Dolomites, late Eocene-early Oligocene volcanic activity to the north of Verona exposed ultramafic to acidic volcanic rocks of upper mantle origin in the “Venetian Volcanic Province” [*Barbieri et al.*, 1978].

## 3. Calculation of Lithospheric Strength Profiles

[9] The construction of lithospheric strength envelopes requires knowledge of the geometry, the petrology and thermal state of the area of interest (see below). Dependent on these data and the prevailing stress regime the critical stress (yield strength) required to cause non-recoverable rock deformation can be calculated for any given depth (for a review, see *Ranalli* [1997]). As a function of the dominant deformation mechanism empirically deduced laws have been formulated describing rock deformation. We use

Byerlee’s law to describe the frictional strength of the lithosphere [*Sibson*, 1974]

$$(\sigma_1 - \sigma_3) = \alpha \rho g z (1 - \lambda), \quad (1)$$

where  $(\sigma_1 - \sigma_3)$  is the critical stress difference,  $\alpha$  is a parameter depending on the type of faulting,  $\rho g z$  is the overburden pressure, and  $\lambda$  the pore fluid factor, which is the ratio of the pore fluid pressure to the overburden pressure; and flow laws such as power law

$$\sigma_1 - \sigma_3 = \left( \frac{\dot{\epsilon}}{A_p} \right)^{\frac{1}{n}} \exp \left[ \frac{E_p}{nRT} \right] \quad (2)$$

and Harper-Dorn creep

$$\sigma_1 - \sigma_3 = \sigma_D \left[ 1 - \sqrt{-\left( \frac{RT}{E_D} \right) \ln \left( \frac{\dot{\epsilon}}{A_D} \right)} \right] \quad (3)$$

for its creep strength [e.g., *Goetze and Evans*, 1979; *Carter and Tsenn*, 1987] where  $(\sigma_1 - \sigma_3)$  is the critical stress difference,  $\dot{\epsilon}$  is the strain rate,  $T$  is the temperature,  $R$  is the gas constant, and  $A_p$ ,  $E_p$ ,  $E_D$ ,  $A_D$  and  $n$  are material dependent quantities. In our calculations the creep strength of olivine in excess of 200 MPa is calculated using the Harper-Dorn creep law.

[10] From equations (1), (2), and (3) it is evident that the frictional strength of the material is strongly pressure dependent whereas the ductile strength is sensitive to temperature and strain rate. For each grid point in the model both frictional as well as ductile strength are calculated denoting the lower value as the yield strength at that particular point. We use a strain rate of  $10^{-14} \text{ s}^{-1}$  representing an average strain rate for a collisional setting such as the Alps [*Pfiffner and Ramsay*, 1982]. In the model we assume a uniform distribution of strain rate yet strain rate may vary considerably across an orogen as shown by numerical modeling [*Pfiffner et al.*, 2000]. The overall present-day stress field in the Eastern Alps is one of roughly N-S compression and E-W extension although it appears to be heterogeneous on a smaller scale due to the superimposition of different stress provinces [*Bressan et al.*, 1998; *Gerner et al.*, 1999; *Reinecker and Lenhardt*, 1999]. The strength along the TRANSALP profile is therefore calculated assuming compressional deformation in N-S direction.

[11] The total strength of the lithosphere is calculated through vertical integration of the yield envelop according to the following relationship:

$$\sigma_L = \int_0^h \sigma_y(z) dz \quad (4)$$

where  $\sigma_L$  is the total lithospheric strength and  $h$  the thickness of the lithosphere.

[12] We regard the modeling predictions as upper bound to the lithospheric strength because the applied



failure criterion (Byerlee's Law) does not take into account possible "high-pressure failure" [Shimada, 1993], and the simultaneous operation of frictional sliding and power law creep in the brittle-ductile transition zone [Handy *et al.*, 1999]. Moreover, the calculations have been performed in the absence of horizontal stresses like far-field stresses or bending stresses, which both cause a reduction of the lithospheric strength [e.g., Cloetingh and Burov, 1996].

#### 4. Geometry and Composition of the Modeled Cross Section

[13] The crustal-scale geometry along the TRANSALP profile is constrained by the results of the deep-seismic reflection campaign comprising explosion and vibroseis data combined with passive seismic experiments [TRANSALP Working Group, 2002] (Figures 2a and 2b). For the strength calculations we used the interpretation of the above quoted data (Figure 2c) as suggested by the TRANSALP Working Group [2002]. Since the differences of the offered interpretations mainly concerns the structure of the Eastern Alps underneath the Periadriatic Fault and to a lesser extent the distribution of upper and lower crustal layers, we perform our strength calculations based on the geometry of the "Lateral Extrusion Model" (Model B of TRANSALP Working Group [2002]) only. Key features of this interpretation comprise crustal-scale ramps along which the Adriatic lower crust has been upthrust and backthrust and the presence of a thin European and thick Adriatic lower crust (Figure 2c). The obtained data, however, do not allow drawing unequivocal conclusions on the Moho and lower crustal geometry at the deepest part of the crustal root.

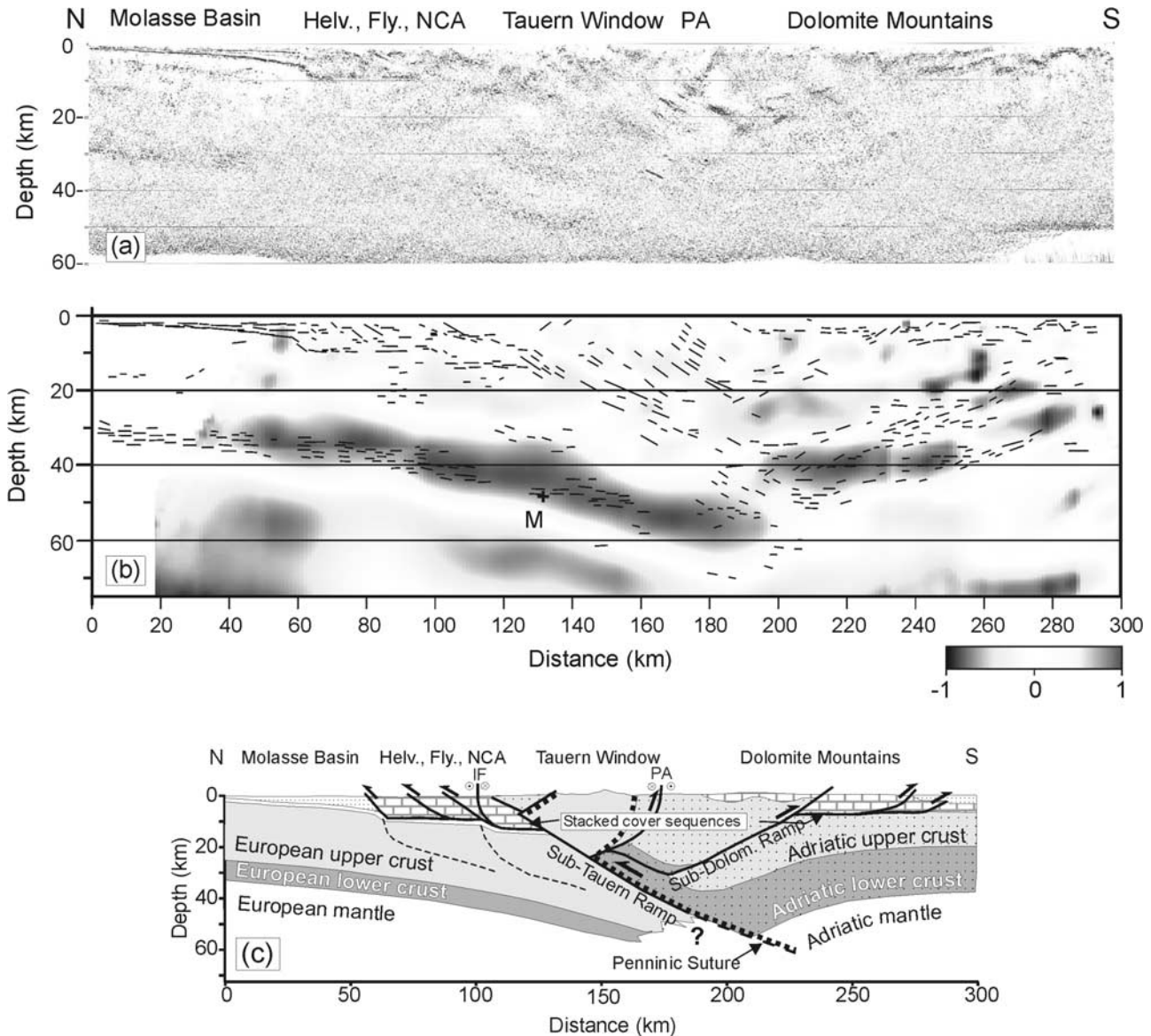
[14] In our simplified rheological model we distinguished (1) the Molasse sediments, (2) the (stacked) cover series (Helvetic, Flysch and northern Calcareous Alps in the north and Dolomites in the south), (3) the crystalline upper crust, (4) the lower crust, and (5) the mantle lithosphere. Subsequently, we assigned rock analogues to the different layers in agreement with the surface geology, conventional rheological models for continental lithosphere and gravity data [e.g., Carter and Tsenn, 1987; Ebbing *et al.*, 2001].

#### 5. Thermal Model Along the TRANSALP Section

[15] Present-day surface heat flow densities derived from temperature measurements in shallow wells (usually not deeper than 4 km) or at engineering sites (e.g., tunnel excavations) are subject to various near-surface processes like ground water circulations, erosion, sedimentation, uplift, subsidence or climate. In deeper levels of the crust or the mantle lithosphere horizontal and vertical movements of rock units [e.g., Mancktelow and Grasemann, 1997] or the emplacement of magmatic bodies induce transient thermal perturbations, which are not easy to

extract from surface heat flow densities. In order to account for the advective heat transport within the lithosphere, the thermal structure along the TRANSALP section was calculated using a kinematic model that solves for the heat transfer equation in two dimensions. A concise description of the numerical model is given by van Wees *et al.* [1992]. Different from the model of Genser *et al.* [1996] this model was set up to specifically fulfill the geometrical boundary conditions deduced from the deep seismic line. Additionally, new constraints on the vertical movements in the Tauern Window and surroundings have been incorporated [Fügenshuh *et al.*, 1997; Stöckhert *et al.*, 1999; Müller *et al.*, 2000; 2001; Liu *et al.*, 2001]. The model is not designed to exactly reproduce the temperature-time (T-t) history at any given point along the TRANSALP section but aims to account for the influence of advective heat transport (either downward or upward) on the thermal structure of the lithosphere during the main deformation phases since the Eocene including underthrusting of the Zentralgneiss and Helvetic Units and associated sedimentation in the fore-deep or exhumation of rocks in the Tauern Window area. Erosion was used as exhumation mechanism throughout the model run. In this study we adopt the exhumation rates calculated by Fügenshuh *et al.* [1997] based on temperature-time data for the western Tauern Window. Following Frisch *et al.* [1998] we applied a total amount of N-S shortening for the Eastern Alps north of the Periadriatic Fault of 110 km for the last 30 Ma. Model limitations do not allow accounting for simultaneous backthrusting in the Southern Alps without disturbing and displacing the thermal structure underneath the Tauern Window. However, model runs neglecting the thermal structure north of the Periadriatic Fault but incorporating thrusting in the Southern Alps by adopting 22 km of shortening for the Valsugana thrust system and 30 km for the Grappa thrust [Schönborn, 1999] yielded a "colder" Adriatic plate than shown in Figure 3. Down to the 700°C isotherm the difference is little (<10%) but increases toward the asthenosphere-lithosphere boundary to up to 20%. Since the temperature difference is fairly small for the thermal range relevant to the strength calculations we use the temperature structure shown in Figure 3. Furthermore, the numerical model does not take into account the affect of topography or the emplacement of plutons on the thermal field. Since topography only affects the shallow part of the crust [e.g., Stüwe *et al.*, 1994] and late Eocene-Oligocene plutonism and volcanism are of little volume, neglecting these processes does not alter the strength calculations significantly.

[16] The predicted thermal structure along the TRANSALP profile (Figure 3) is calculated for the thermal parameters and boundary conditions described in the caption to Figure 3 and listed in Table 1 and retains the memory of the underthrusting European plate along a major thrust and the rapidly exhuming Zentralgneiss unit in the Tauern Window. The former causes a downward deflection of the isotherms in the area of the Molasse Basin, the Northern Calcareous Alps and the Southern Alps, whereas the latter



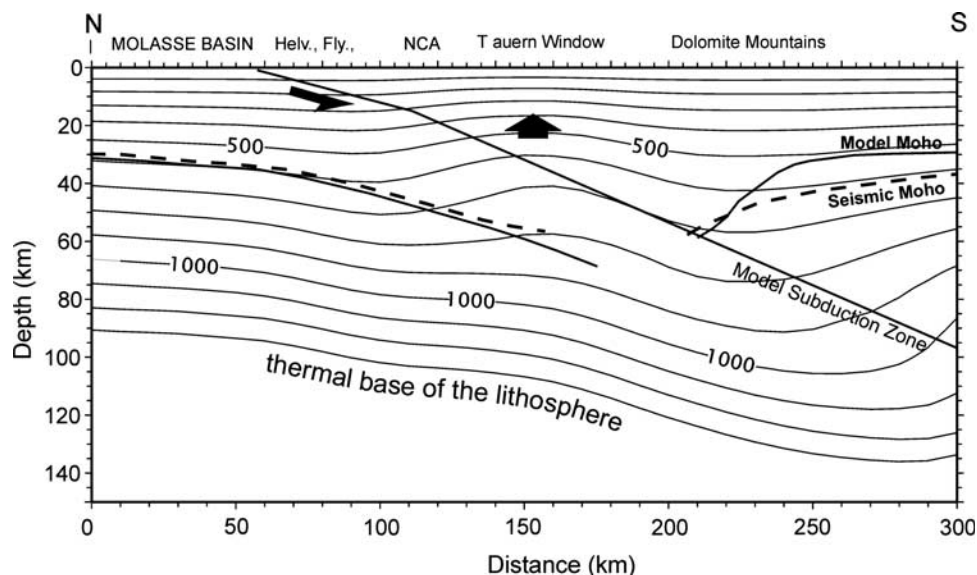
**Figure 2.** Sections along the TRANSALP transect showing (a) migrated reflection seismic data (vibroseis data); (b) line drawing of reflections deduced from vibroseis and explosive sections superimposed on depth-migrated receiver functions, and (c) simplified interpretation of the seismic data. All figures are taken from *TRANSALP Working Group* [2002]. Note that Figures 2b and 2c have been modified slightly. The large scale structure at the Inntal Fault has been adopted after *Reiter et al.* [2003]. The cross in Figure 2b marks the position of the Moho (M) deduced from refraction seismic data along the ALP75 section [Yan and Mechie, 1989]. Fly, Flysch Zone; Helv., Helvetic Zone; IF, Inntal Fault; NCA, Northern Calcareous Alps; PA, Periadriatic Fault. See color version of this figure at back of this issue.

causes an upward deflection of the isotherms in the Tauern Window area. While the influence of the Tauern Window exhumation decreases with depth, that of the underthrusting event increases with depth but has little effect on the shallow part (down to 500°C isotherm) of the Adriatic plate. Compared with recently compiled surface heat flow data [Sachsenhofer, 2001; Della Vedova et al., 2001], our predicted surface heat flows are higher by 20% on average. However, inferences from our model calculations are consistent with heat flows deduced by Della Vedova et al.

[2001] for the Southern Alps when they correct the surface heat flow data for near surface processes.

## 6. Lithospheric Strength Along the TRANSALP Transect

[17] In the following section we present model predictions for two end-member scenarios. Model A (MA) represents a mechanically strong lithosphere, whereas Model B



**Figure 3.** Present-day thermal model for the TRANSALP section calculated with a two-dimensional explicit 3-step Runge-Kutta finite difference model [van Wees *et al.*, 1992]. The left-hand and right-hand sides of the numerical model are fixed ( $\delta T/\delta x = 0$ ) during time integration. Material and thermal parameters used for the thermal modeling are listed in Tables 2 and 3. Simple shear, pure shear, as well as combined shear tectonics are simulated based on a velocity field approach [e.g., van Wees *et al.*, 1992]. The rates of tectonic movements are constant for individual time steps. Blocks, separated by faults, have the ability to move with different velocities, parallel to the fault plane. Temperatures are displayed as isotherms with a spacing of 100°C. Note that the 1300°C isotherm represents the lithosphere-asthenosphere boundary in the thermal model, which is allowed to move according to the imposed tectonic movements. We follow Pollack and Chapman [1977] and assume that 40% of the surface heat flow is accounted for by radiogenic heat production within the upper crust. Underthrusting of the European plate was accommodated along a thrust, which largely coincides with the location of the Sub-Tauern Ramp in the deeper parts of the section and with the base of the stacked cover units underneath the Northern Calcareous Alps, the Flysch Zone and the Helvetic zone, respectively. Note that the “Seismic Moho” of the Adriatic plate in Figure 3 is not as well constrained as the European Moho. Abbreviations as Figure 2.

(MB) stands for a weak lithosphere. In order not to lose valuable information on subtle strength variations within the low strength regions we only display stresses up to 300 MPa. Rheological parameters are listed in Tables 2 and 3.

[18] Model A is characterized by the presence of three mechanically strong layers comprising the upper and lower crust and the upper mantle (Figure 4a). Lateral variations of strength within the upper crust are due to lateral variations of the temperature structure reflected in the decreasing

strength of the upper crust in the Tauern Window area and lateral changes in composition such as the change from Molasse sediments to bedrock and the change of the stacked cover sequences (Northern Calcareous Alps, Flysch), which we ascribed a quartz rheology, to crystalline upper crust (e.g., at profile km 100–130). The strength of this horizon denotes it as potential detachment surface allowing for decoupling of the cover nappes from the European crystalline basement. Mainly because of its deeper position where

**Table 1.** Material Parameters for Thermal Modeling<sup>a</sup>

Layer	Thickness, km	Density $\rho$ , kg m <sup>-3</sup>	Conductivity $\kappa$ , W m <sup>-1</sup> °C <sup>-1</sup>	Specific Heat $c_p$ , J kg <sup>-1</sup> °C <sup>-1</sup>	Heat Production A, $\mu$ W m <sup>-3</sup>
Upper crust	20	(2700)	(2.7)	1050	(2.0)
Lower crust	10	2900	2.5	1050	0.5
Mantle		3300	3.1	1050	0

<sup>a</sup>Thermal parameters are after Pollack and Chapman [1977], and those that are after Vosteen *et al.* [2001] are in parentheses. Model dimensions are 1600 × 250 km. Initial surface temperature/surface heat flow is 0°C/60 mWm<sup>-2</sup>. Grid spacing is 4.5 × 2.5 km. Temperature and initial heat flow at the asthenosphere-lithosphere boundary is 1300°C/31 mWm<sup>-2</sup>.



**Table 2.** Rheological Parameters Used for Strength Calculations of Models A and B<sup>a</sup>

Layer	Minerals/Rock	Power, n	Activation Energy, Q, Jmol <sup>-1</sup>	Initial Constant, Pa <sup>-N<sub>s</sub>-1</sup>	Model
Cover nappes	quartzite dry	1.9	172600	1.26e-13	A
	quartzite wet	2.72	134000	6.03e-24	B
Upper crust	quartzite dry	1.9	140600	7.94e-16	A
	quartzite wet	3.3	186500	3.16e-26	B
Lower crust	mafic granulite	2.4	212000	1.26e-16	A
	diorite wet	4.2	445000	8.83e-22	B
Mantle	olivine dry	3.0	510000	7.00e-14	A, B

<sup>a</sup>Rheological parameters are after *Carter and Tsenn* [1987] and *Goetze and Evans* [1979]. Acceleration of gravity (g) is 9.81 ms<sup>-2</sup>. Universal gas constant (R) is 8.314 J mol K<sup>-1</sup>. Static friction coefficient (fs) is 0.6. Hydrostatic pore fluid factor ( $\rho_w/\rho$ ) is ~0.35.

temperatures are higher, the European lower crust is predicted to be weaker than the Adriatic lower crust. A local strength maximum at the top part of the upthrust Adriatic lower crust is related to its shallow position. With depth, the strength of this layer decreases to reach a minimum where the dip of the layer changes (profile km 180–190). Southward the strength of the Adriatic lower crust, which is influenced by the cooling effect of the underthrust European plate increases again. For the same reason the Adriatic upper mantle is predicted to be stronger than that of the European plate. Coupling among layers of the Adriatic plate is predicted to be strong whereas decoupling of the European upper crust from the coupled lower crust and upper mantle is suggested. For MA the integrated strength of the whole lithospheric is calculated to be highest for the Adriatic plate and least for the Tauern Window region (Figure 4a, top panel). Except for the central part of the orogen the upper mantle strength contributes most to the whole lithospheric strength along the TRANSALP section. The strength of the upper crust is most important in the central weak zone, where it largely controls the bulk strength of the lithosphere. In contrast to the Adriatic lower crust, the European lower crust adds little to total lithospheric strength.

[19] Different to MA the model lithosphere of MB consists of two strong layers only, of which the thickness of the strong upper crust is distinctly reduced (Figure 4b). MB suggests strong decoupling of the mechanical strong part of the upper crust along a thick layer comprising the ductile part of the upper crust and the entire lower crust from the underlying mantle lithosphere. Similar to MA least strength is predicted for the central part of the orogen. The lack of lower crustal strength affects the total strength of the Adriatic plate most, which is nearly equally strong as the European lithosphere in MB. (Figure 4b, top panel).

## 7. Strength Models and Seismicity Along the TRANSALP Transect

[20] Seismicity is commonly regarded as an expression of frictional sliding along discrete faults under brittle deformation conditions, thus providing information on the depth extent and the distribution of brittle layers in the lithosphere

[e.g., *Scholz*, 1988]. We projected the earthquake hypocenters along a 50 km wide zone with the TRANSALP profile in its center onto the interpretation of the deep seismic line together with lithospheric strength envelopes at selected sites to infer possible correlations between the depth extent of the brittle layers and the cut-off depth of seismicity (Figure 5). Since we did not explicitly calculate the strength in the brittle-ductile transition by applying a coupled frictional-viscous flow law [*Handy et al.*, 1999], we regard the calculated brittle-ductile boundary as a first-order approximation to a qualitative understanding of the distribution of brittle layers along the TRANSALP section. Although, the seismicity distribution of historical earthquakes does not differ from the instrumentally recorded earthquakes significantly, we display instrumental earthquakes, only, for reasons of having the most accurate available depth information for the seismic events. Plotted are seismic events for the time periods from 1971–2000 for Southern Germany and Austria and from 1977–2000 for Northern Italy, respectively. Data have been provided by the Central Institute for Meteorology and Geodynamics (ZMAG) in Vienna and by the Istituto Nazionale di Oceanografia e di Geofisica Sperimentale in Trieste, published online in the bulletin to the Friuli Venezia Giulia Seismometric Network.

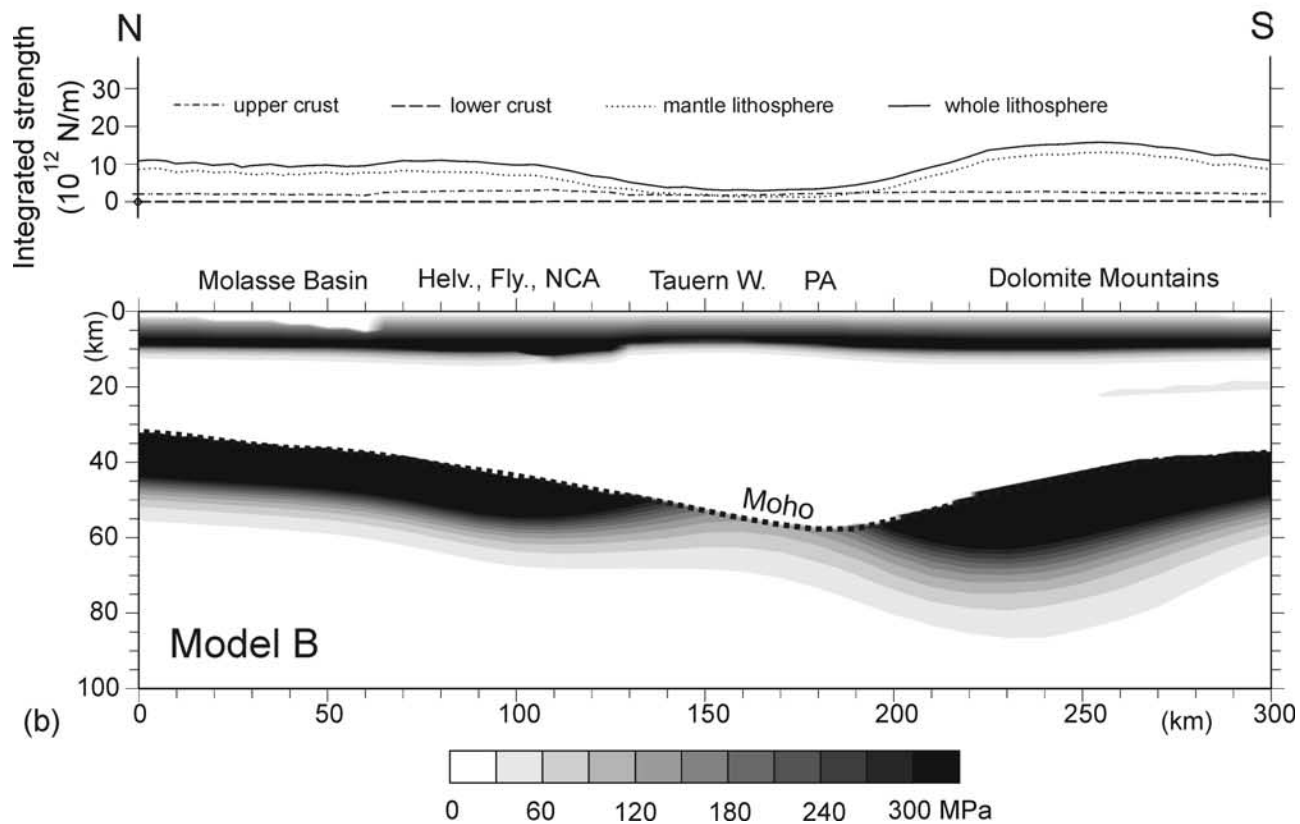
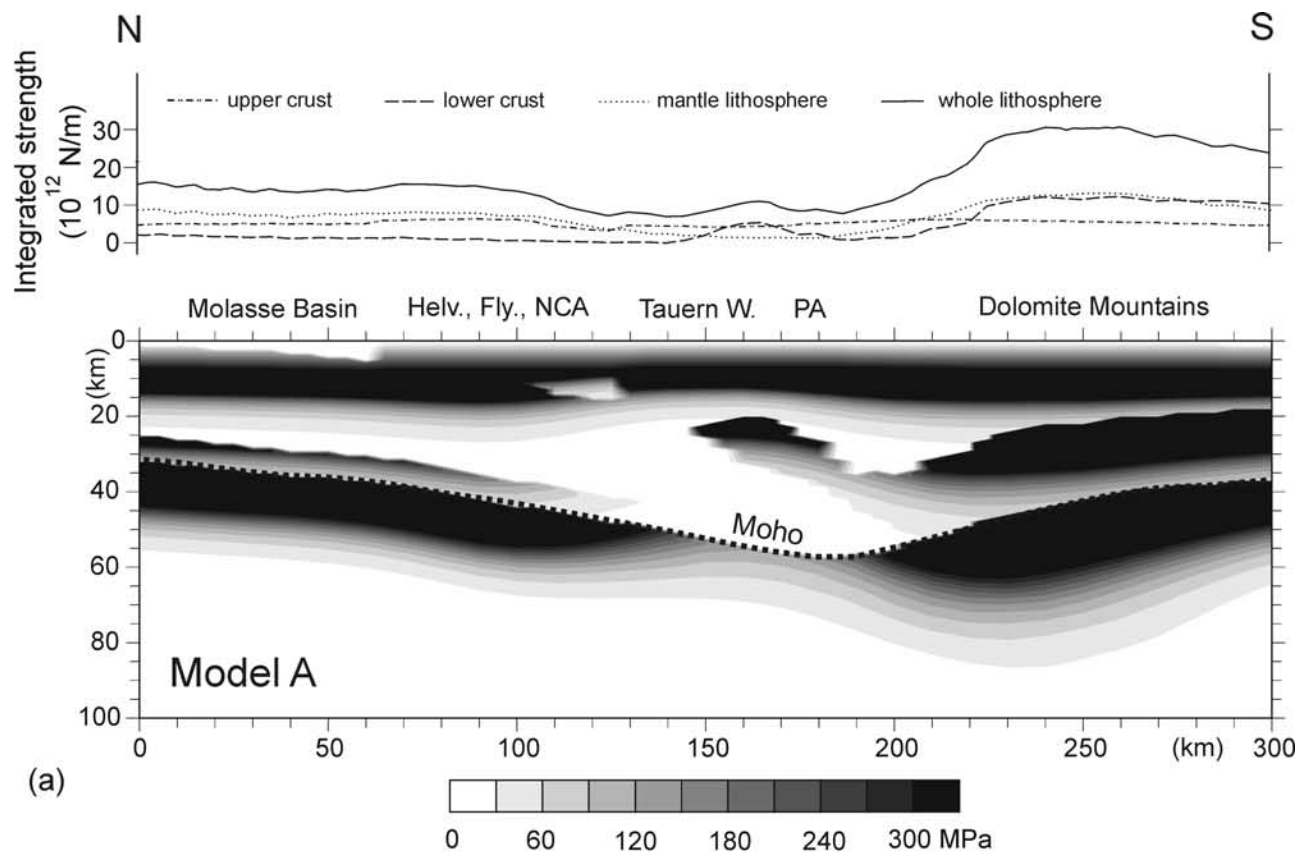
[21] For the considered time period all earthquakes along the investigated sector nucleated above the Moho. Striking is the low seismicity within the European plate and a strong concentration of seismicity within the Adriatic plate owing to an intense fragmentation and rotation of the Adriatic plate at the junction to the Dinarides [*Slejko et al.*, 1989; *Bressan et al.*, 1998].

[22] For MA all the focal depths to the north of the Tauern Window are shallower than the predicted brittle-ductile boundary (12–14 km) and retain within the upper

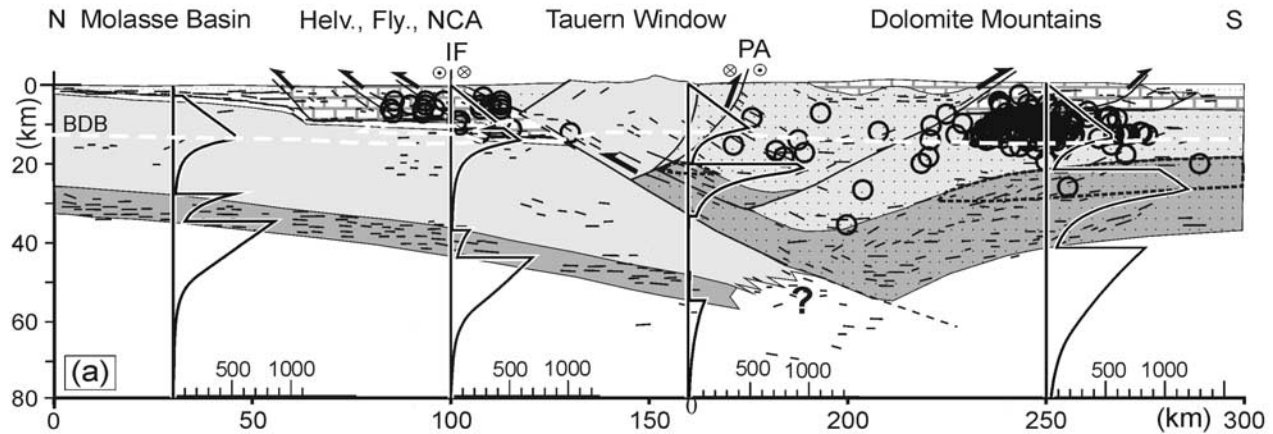
**Table 3.** Harper-Dorn Creep Flow Parameters for Dry Olivine Used in Models A and B

	Value
$\sigma_D$ , Mpa	$8.5 \times 10^3$
$E_D$ , kJmol <sup>-1</sup>	535
$A_D$ , s <sup>-1</sup>	$5.7 \times 10^{11}$

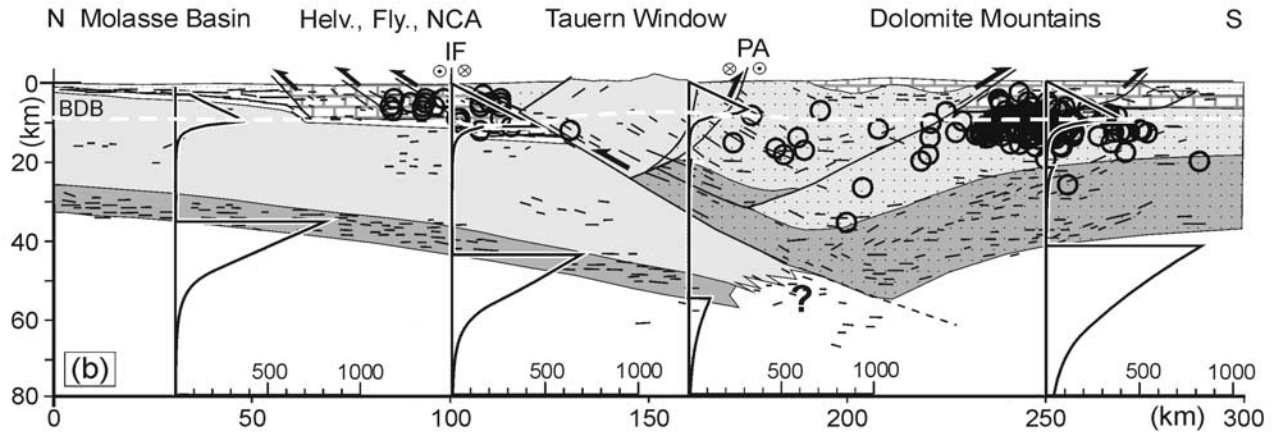




## Model A



## Model B

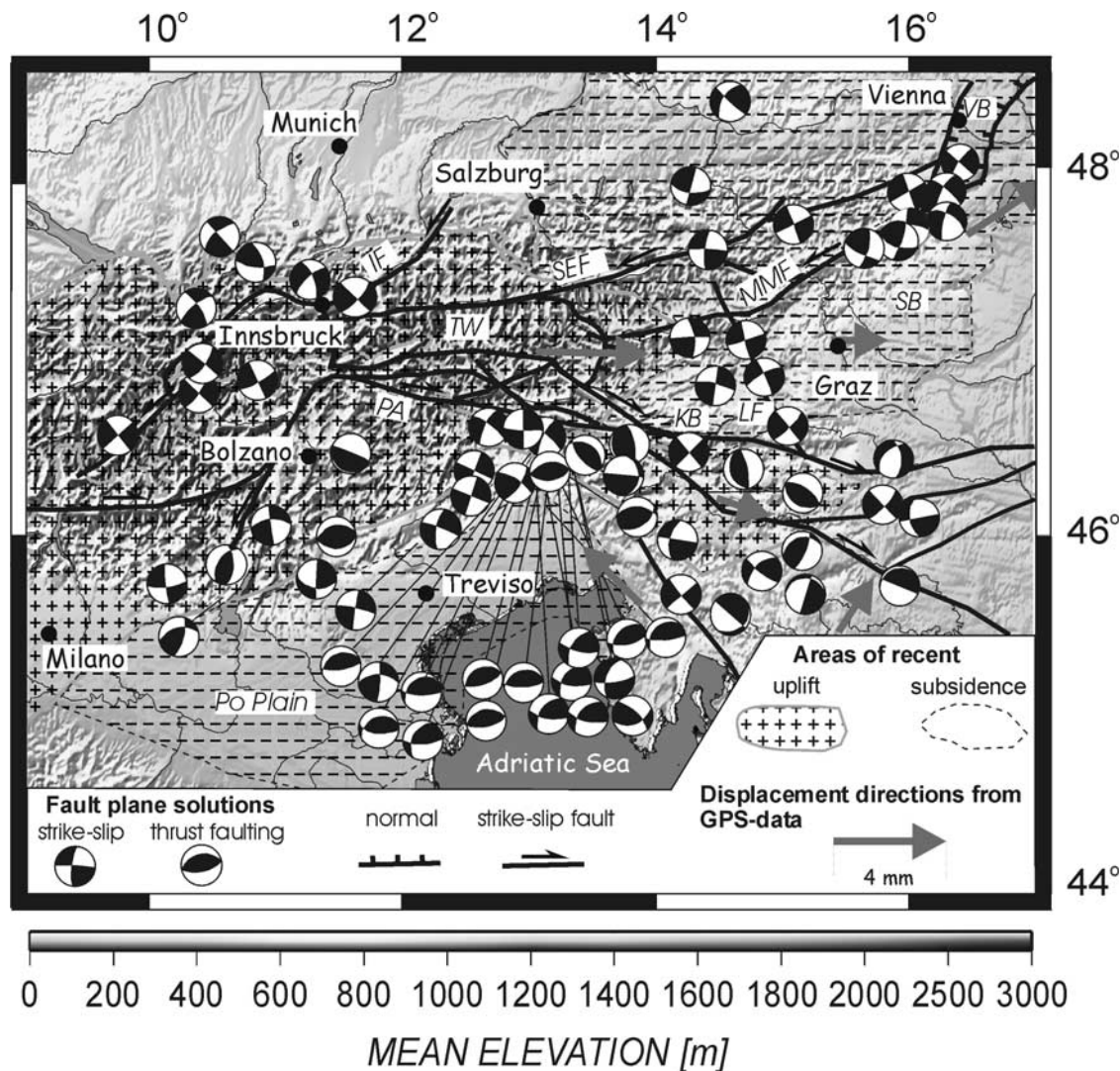


**Figure 5.** Strength envelopes (in MPa) and earthquake hypocenters superimposed on the interpreted TRANSALP Line for (a) Model A and (b) Model B. BDB denotes the brittle-ductile-boundary as indicated by the dashed white line. Note that the Penninic Suture has not been taken into account in the modeling. The lateral spread of seismicity data along the Inn Valley and other faults is due to projecting the seismicity of obliquely trending structures onto the TRANSALP line. Stippled lines in Figure 5a delineate the brittle parts of the Adriatic lower crust. Color coding as Figure 2c and abbreviations as Figures 2a and 2c.

12 km of the brittle crust (Figure 5a). This seismicity is mainly related to left-lateral movements along the Inntal Fault system and to the reactivation of preexisting extensional structures within the uppermost part of the European crystalline basement [Reiter *et al.*, 2003]. Although, a shift of the hypocenters to greater depths can be observed to the south of the Tauern Window, still about 95% of the earthquakes plot within the predicted brittle cores of the upper and lower crust, respectively. Noteworthy is that the seismic activity within the predicted ductile parts of the Adriatic

upper crust, which might be an expression of high strain rates within shear zones, occurs within its mechanical strong part, yet the rheology is predicted to be ductile. As it has also been shown in previous studies [e.g., Okaya *et al.*, 1996] lateral variations of strain rate influence the depth extent of the brittle layers. Other explanations for seismicity within the ductile crust comprise the presence of plastic instabilities [e.g., Ord and Hobbs, 1989] or by reducing the frictional resistance to brittle failure through the presence of fluids [e.g., Deichmann, 1992].

**Figure 4.** (opposite) Strength predictions for (a) Model A and (b) Model b. Top panels show the lateral variation of the integrated strength in  $10^{12} \text{ Nm}^{-1}$  of the total lithosphere and its layers. Abbreviations as Figure 2.



**Figure 6.** Summary of neotectonic data of the Eastern Alps. Data are taken from *Arca and Beretta* [1985], *Slejko et al.* [1989], *Kahle et al.* [1997], *Bressan et al.* [1998], *Gerner et al.* [1999], *Reinecker and Lenhardt* [1999], *Grenerczy et al.* [2000], and *Höggerl* [2002]. Note that for the seismically very active regions of northern Italy, Slovenia, and Croatia only a representative selection of focal mechanism can be shown. IF, Inntal Fault; KB, Klagenfurt Basin; LF, Lavanttal Fault; Mur, Mürz Fault; PA, Periadriatic Fault; SB, Styrian Basin; SEF, Salzach-Ennstal Fault; TW, Tauern Window; VB, Vienna Basin. See color version of this figure at back of this issue.

[23] For MB the correlation of the seismicity with the distribution of brittle layers is generally poor (Figure 5b). For this model a depth of the brittle-ductile boundary between 7 and 11 km is predicted. To the north of the Inntal Fault the calculated brittle-ductile boundary follows approximately the base of the Northern Calcareous Alps, which is considered as important detachment level [e.g., *Auer and Eisbacher*, 2003] and defines the lower bound of the seismogenic layer. Between the Inntal Fault and the Tauern Window the earthquake hypocenters extend below the predicted depth of the brittle ductile boundary. To the south of the Tauern Window only about 40% of the seismicity correlates positively with a shallow brittle ductile

transition suggesting that MB underestimates the strength of the Adriatic Plate.

## 8. Implications for the Neotectonics of the Eastern and Southern Alps

### 8.1. Neotectonic Data for the Eastern and Southern Alps: A Synthesis

[24] Although the evaluation of seismic activity of regions or structures may be biased by the short observation period, some general conclusions can be drawn from the distribution of earthquake epicenters (Figure 6) in the study



area: (1) The European foreland of the Eastern Alps including the Bohemian massif is marked by relative modest seismic activity with focal depths down to ca. 12 km [Reinecker and Lenhardt, 1999; G. Grünthal, personal communication, 2000], whereas earthquakes in the foreland of the central and the Western Alps are more frequent and occur within the entire crust [Deichmann, 1992; Pavoni *et al.*, 1997; Schmid and Kissling, 2000]. (2) Most of the earthquakes within the orogenic wedge of the Eastern Alps occur within the upper 13 km of the crust and along major faults inherited from Miocene extrusion tectonics. High seismicity is observed for the Vienna Basin, the Mur-Mürz Fault, the Salzach-Ennstal Fault, the Lavanttal Fault or the Inntal Fault. Fault plane solutions calculated for seismic events point to a strike-slip deformation regime (Figure 6). (3) To the south of the Periadriatic Fault, which itself shows fairly little seismic activity, the seismicity increases dramatically and reaches a climax in the Friuli area at the junction of the eastern Southern Alps and the Dinarides. Fault plane solutions indicate strike-slip and thrust faulting regimes [Slejko *et al.*, 1989; Bressan *et al.*, 1998; Gerner *et al.*, 1999] (Figure 6). Most of the hypocenters are in a depth range of 4–15 km with a concentration between 8 and 13 km [Bressan *et al.*, 1998].

[25] Vertical crustal movements deduced from levelling studies [Arca and Beretta, 1985; Kahle *et al.*, 1997; Höggerl, 2002] indicate relative uplift of the central and Eastern Alps with respect to the European foreland, the Adriatic foreland, the Eastern Alps east of the Tauern Window, and the Styrian Basin (Figure 6). Highest uplift rates (up to 1.6 mm/year have been inferred in Austria for the Tauern Window region and in eastern Switzerland for the regions of the Penninic nappes to the southwest of the Engadine Window. Uplift rates decay gradually away from these regions. In the Eastern Alps of Austria the change of relative uplift to relative subsidence occurs east of the Tauern Window, where the Moho shallows rapidly from about 50 km to 35 km [Miller *et al.*, 1978]. Relative subsidence is fastest in the eastern Styrian Basin, the Vienna, and Klagenfurt Basins and a zone along the Lavanttal. Relative uplift to the south of the Periadriatic Fault decreases southward and turns into relative subsidence at the transition to the Po Plain, where subsidence is highest in the Po delta region [Arca and Beretta, 1985; Slejko *et al.*, 1989; Carminati and Di Donato, 1999]. Relative uplift of the Southern Alps at the transition to the Po plain (Montello region) appears to be related to active thrusting along blind thrusts in the subsurface [Benedetti *et al.*, 2000].

[26] First results of geodetic projects focussing on the horizontal crustal movements making use of the Global Positioning System (GPS) emphasize eastward motion of the central Eastern Alps guided by major faults namely the Mur-Mürz Fault in the north and the Periadriatic Fault in the south (Figure 6). Maximum horizontal velocities calculated for a station at the eastern margin of the Tauern Window are in the order of 2–4 mm/year relative to stable Europe [Grenerczy *et al.*, 2000]. South of the Periadriatic Fault movements are dextral with a N-S component of shortening in the order of 2 mm/year. The direction of

movements within that strongly faulted region appears to be fault block dependent [Grenerczy *et al.*, 2000, Figure 7].

[27] From the above quoted data together with analysis of borehole break-outs and in-situ stress measurements a bulk N-S directed orientation of the maximum compressive stress has been inferred for the Eastern and Southern Alps whereas the Dinarides and the eastern part of the Pannonian Basin are subject to NE-SW directed compression [Bressan *et al.*, 1998; Gerner *et al.*, 1999; Reinecker and Lenhardt, 1999]. Both stress directions are related to the northward motion and counterclockwise rotation of the Adriatic microplate [e.g., Bressan *et al.*, 1998; Gerner *et al.*, 1999; Caporali and Martin, 2000].

## 8.2. Lateral Variations of Lithospheric Strength and its Relation to Neotectonics

[28] The importance of lateral variations of lithospheric strength for the post-Late Oligocene evolution of the Eastern Alps has been demonstrated by Ratschbacher *et al.* [1991b] through analogue modeling aiming to study the boundary conditions for lateral extrusion tectonics. In N-S section their initial setup is very similar to our model prediction (MA) in as far as a thick and weak central part, representing the orogen, is flanked on either side by stronger units, representing the European foreland and the Adriatic indenter, respectively (Figure 4a). Upon compression such a system reacts by a combination of thickening in front of the indenter and lateral escape of material in E-W direction, preconditioned by a weak lateral confinement. In the case of the Eastern Alps, N-S contraction caused thrusting, folding and strike-slip deformation in the area of maximum horizontal compression (Tauern Window) passing laterally into transtensional and normal fault structures related to spreading and escape [Ratschbacher *et al.*, 1991a, 1991b; Neubauer *et al.*, 2000a]. Escape toward the Pannonian-Carpathian region probably was facilitated by slab-pull forces arising from the subduction of oceanic or thin continental crust along the Carpathian arc [e.g., Horváth, 1993].

[29] Although the stress sources have changed since the climax of extrusion in the Eastern Alps, subduction along the Carpathian arc and concomitant extension in the Pannonian Basin [Bada *et al.*, 2001], we will argue below that the relative strength relationships among the major tectonic units prevailed up to now.

### 8.2.1. European Plate

[30] The strength of the European plate underneath and in front of the Eastern Alps, parameterized by the effective elastic thickness (EET) of the downbending plate, has been inferred from forward modeling of gravity and topographic data [Stewart and Watts, 1997] and flexural modeling studies [e.g., Andeweg and Cloetingh, 1998, and references therein]. Both approaches yielded EET values in the range of 20–30 km for profiles close to the TRANSALP transect denoting the European plate as being strong. These values are consistent with strength estimates for the Bohemian Massif [Lankreijer *et al.*, 1999] and our strong model set up (MA), for which we estimate the EET to be in the order of 23–27 km using the equation of Burov and Diament [1995]

for decoupled layers and a critical strength value of 50 MPa for the definition of the thickness of the mechanically strong parts of the layers. Interestingly, *Pfiffner et al.* [2002] arrived at comparable EET estimates ( $\sim 25$  km) for the early loading phase ( $\sim 32$  Ma ago) of the European plate underneath the mountain range suggesting that the time affects on the bulk strength of the European plate are fairly low. The relative high strength of the European plate and in particular that of the Bohemian Massif predicted by numerical modeling is in accordance with its Variscan thermo-tectonic age [*Cloetingh and Burov*, 1996] and finds support by geodetic data, showing very little horizontal and vertical movements, and by the scarcity of seismic events, which occur mainly at shallow crustal levels ( $< 10$  km) [*Reinecker and Lenhardt*, 1999]. The few seismic events extending into the crystalline basement of the European plate (Figure 5) along the TRANSALP section also favors a “strong European plate model” and probably suggests that the stress level within the crust rarely reaches the yield limit.

### 8.2.2. East Alpine Orogenic Wedge

[31] As also suggested for the geologic past, prior to the onset of extrusion [*Ratschbacher et al.*, 1991b], the present-day crustal thickness decreases from west to east along the central axes of the Eastern Alps [*Miller et al.*, 1978]. The thickness variation appears to reflect the natural termination of the Zentralgneiss unit (no equivalent continental block in the Rechnitz Window area) and the influence of orogen parallel extension. Rapid exhumation of medium-grade metamorphic units in the Tauern Window exerts a strong control on the temperature structure and subsequently the strength of the crust. Correspondingly, the predicted lithospheric strength is lowest for the central segment of the TRANSALP profile agreeing with low EET's inferred for thermo-tectonically young lithosphere [*Cloetingh and Burov*, 1996]. Our results also confirm findings of *Genser et al.* [1996] and *Okaya et al.* [1996] who postulate, based on thermo-mechanical modeling, weak lithosphere for the Eastern Alps across the Tauern Window and the central part of the Swiss central Alps, respectively. To the north of the Tauern Window seismicity is confined to the upper 12 km of the crust and is mainly related to preexisting weakness zones like the Inntal Fault or the basement-cover interface (Figures 5 and 6) along which the yielding is reached at low stress levels. Interestingly, there is little seismic activity in the Tauern Window area itself suggesting that stresses are dominantly released by aseismic creep. In the vicinity of the Periadriatic Fault, shallow seismic events are rare and seismicity occurs down to depths of ca. 16 km (Figure 5). Along strike, the predicted weakness of the central segment coincides with the extrusion corridor within which present-day and paleo-heatflows have been higher relative to the surrounding units [*Sachsenhofer*, 2001]. Present-day surface heat flow densities up to  $145 \text{ mWm}^{-2}$  in the Styrian Basin and other parts of the Pannonian Basin cause a significant reduction of lithospheric strength resulting in very low EET values of ca. 5–7 km [*Sachsenhofer et al.*, 1997; *Lankreijer et al.*, 1999, and references therein].

[32] Furthermore, areas of maximum uplift (Tauern Window) correspond to the predicted weakest part of the system.

Such a correlation has also been deduced by *Ratschbacher et al.* [1991b] by analogue modeling. In this model the change in height is primarily compression related, whereas other processes such as isostatic adjustment in relation to mountain building, erosion and deglaciation of Alpine glaciers most likely contribute to the total uplift of the Alps as well.

[33] The emerging picture that the central segment of the Eastern Alps is weaker relative to the Adriatic indenter and the European foreland is consistent with GPS data arguing for eastward escape of the central Eastern Alps guided by major strike-slip faults (Figure 6) [*Grenerczy et al.*, 2000]. It is the motion along these faults, which is the major source for the recent seismicity in the Eastern Alps. Eastward escape of the central Eastern Alps is not anymore taken up by shortening along the Carpathian front, which is locked, but by smaller-scale structures within the western Pannonian Basin that are seismically active.

### 8.2.3. Adriatic Indenter

[34] The Adriatic indenter is in our study predicted to be stronger than the central Eastern Alps to the north what is also expressed in the higher EET values ranging from 22 km at the Periadriatic fault to 40 km at profile km 240, and to 31 km at the end of the TRANSALP line. Except for the strong part mentioned above, these values are in line with those (26–30 km) deduced by *Stewart and Watts* [1997].

[35] The same strength relationship must have existed in the geological past to enable shortening and lateral extrusion in the Eastern Alps in front of the north to northwestward protruding indenter [*Ratschbacher et al.*, 1991a]. The main cause for the unequal mechanical behavior arises from the different thermo-tectonic ages (Tertiary versus Triassic). The greater depth extent of earthquake hypocenters appears to support his hypothesis (Figure 5). The majority of the seismic events on the Adriatic plate are related to the movements at the actual plate boundary in the Friuli region (Figure 6). In this region the Adriatic upper crust is strongly fragmented and the seismicity is mainly related to strike-slip and reverse faulting. Most of the earthquakes nucleate at or close to the basement-cover interface in depths between 8 and 12 km [*Slejko et al.*, 1989]. Deeper earthquakes along the TRANSALP section appear to roughly follow the upper-lower crustal boundary where rheological contrasts are high (Figure 5a). In contrast to the Sub-Dolomite Ramp, there is no seismic activity recorded for the deep parts of the Sub-Tauern Ramp, where the Adriatic lower crust is interpreted to have been upthrust to depths of about 20 km (Figure 5a). All the seismicity of that region retains in the upper crust and might coincide with the interface of Zentralgneiss and Adriatic units as shown in the “Crocodile Model” by the *TRANSALP Working Group* [2002]. The intense seismic activity suggests that uplift in the Southern Alps of the Adriatic plate is related to active (back)thrusting along north to NW dipping thrust planes.

### 8.3. Nature of the Lower Crust Along the TRANSALP Transect

[36] The position and geometry of the lower crust along the TRANSALP line has been inferred from the reflectivity

pattern of vibroseis and explosion seismic data (Figure 2). Striking in the interpretation of the *TRANSALP Working Group* [2002] are large differences in thickness between the European and Adriatic lower crusts and the shallow position of the Adriatic lower crust in the hanging wall of the Sub-Tauern Ramp, displaying the geometry of a huge synform. In contrast to the Adriatic lower crust no seismic activity can be associated with the European lower crust along the TRANSALP transect. Such aseismic behavior is also suggested by our model calculations, which do not predict brittle strength for the European lower crust neither for MA nor MB. Similar results have been obtained by *Genster et al.* [1996] for the Eastern Alps and *Okaya et al.* [1996] for the central Alps.

[37] Farther to the west, the wedge shape geometry of the Adriatic lower crust and subduction of European lower crust as suggested for the central Alps, is interpreted to be indicative for a strong lower crust [*Schmid et al.*, 1996; *Pfiffner et al.*, 2000]. In the interpretation of the TRANSALP deep seismic data the European lower crust remains equally thick as far as it can be traced. To maintain this shape the lower crust is not allowed to deform by thickening or thinning arguing for being a strong layer. Dynamic modeling of collision and indentation tectonics [*Willingshofer and Gerbault*, 2002] suggest that relative strength differences of about 30 MPa might be sufficient to concentrate deformation in the weak zones (ductile part of the European upper crust) and to cause little deformation in the stronger layer (lower crust). According to our model calculations such conditions are only fulfilled up to the southern limit of the Northern Calcareous Alps (ca. profile km 130 in Figure 4a). Farther to the south the lower crust is predicted to be weak and hence deformable possibly resulting in its thickening upon compression. If this hypothesis is correct then the reflections in the deep parts of the European upper crust (profile km 120–140) might actually represent the upper limit of a thickened European lower crust similar to a situation as envisaged in the ECORS-CROP cross section through the Western Alps [*Roure et al.*, 1990].

[38] According to the *TRANSALP Working Group* [2002], a key feature in the TRANSALP transect is the major ramp structure (Sub-Tauern Ramp) along which the Adriatic lower crust has been upthrust (Figures 2c and 5a). Critical to this interpretation and its mechanical stability appears to be the strength of the European upper crust, which has to support the weight of the (1) presumably denser Adriatic lower crust and (2) the thickened upper crust above (Figures 5a and 5b, profile km 140–180). Our rheology predictions indicate that the strength of the European upper crust on its own is probably unable to support this “load” (Figure 5a; profile km 160) leading to a gravitationally unstable situation and deformation of the European upper crust in the root zone of the Eastern Alps by viscous flow. Possible sinking of the denser Adriatic lower crust into the less dense underlying European upper crust appears to be prevented by the prevailing N-S to NW-SE contraction of the Eastern Alps. As to whether this contraction still causes upward movement of the Adriatic lower

crust along the Sub-Tauern Ramp remains unknown since the lack of seismicity along this structure is not conclusive concerning its activity. Comparison with analogue experiments investigating accommodation structures in the hanging wall of thrust ramps [*Bonini et al.*, 2000] suggests that at least the frontal part of the Adriatic lower crust probably behaved in a ductile manner when it moved over the Sub-Tauern Ramp. Otherwise (under thoroughly brittle conditions), dissection of the lower crust by numerous backthrusts would have been the consequence (for comparison, see Figure 3 of *Bonini et al.* [2000]).

[39] An important implication of both interpretations of the TRANSALP line [*TRANSALP Working Group*, 2002] for the formation of the suggested crustal-scale geometry is full decoupling of the Adriatic crust from the underlying mantle lithosphere (see Figure 2c). Our strength profiles suggest crust-mantle decoupling only for the model with a weak lower crust (MB), which however fails to explain the seismicity in the upper part of the Adriatic lower crust.

[40] Alternative interpretations of the TRANSALP section supported by high-resolution tele-seismic tomography [*Lippitsch et al.*, 2003] incorporate subduction of the Adriatic plate underneath the orogenic wedge and the European plate [*Schmid et al.*, 2003]. This change in subduction polarity will affect the strength of the different rheological layers: (1) depending on the in depth position of the interpreted boundaries between the upper and lower crust and the mantle lithosphere and (2) possible changes in the thermal structure due to northward subduction of the Adriatic plate. The latter is thought to cause cooling and, hence, strengthening of the European plate. On the scale of the lithosphere the relative strength relations among the major tectonic units described in the previous sections, however, is not expected to change.

## 9. Conclusions

[41] Lithospheric strength predictions for the present-day configuration of the Eastern Alps along the TRANSALP profile suggest that a weak central zone representing the thickest part of the Eastern Alps is flanked to the north and the south by regions of higher strength of the European and Adriatic plates, respectively. Models with weak upper and lower crustal rheologies lead to strong crust-mantle decoupling and the presence of a very thick weak zone extending from the ductile upper crust down to the Moho, whereas a configuration with a strong crust results in stronger crust-mantle coupling and denotes the lower crust as additional load bearing layer. The latter setup is in better agreement with neotectonic and geodetic data, which emphasize that the process of lateral extrusion of fault-bound units from the central zone of the Eastern Alps toward the Pannonian Basin is still active.

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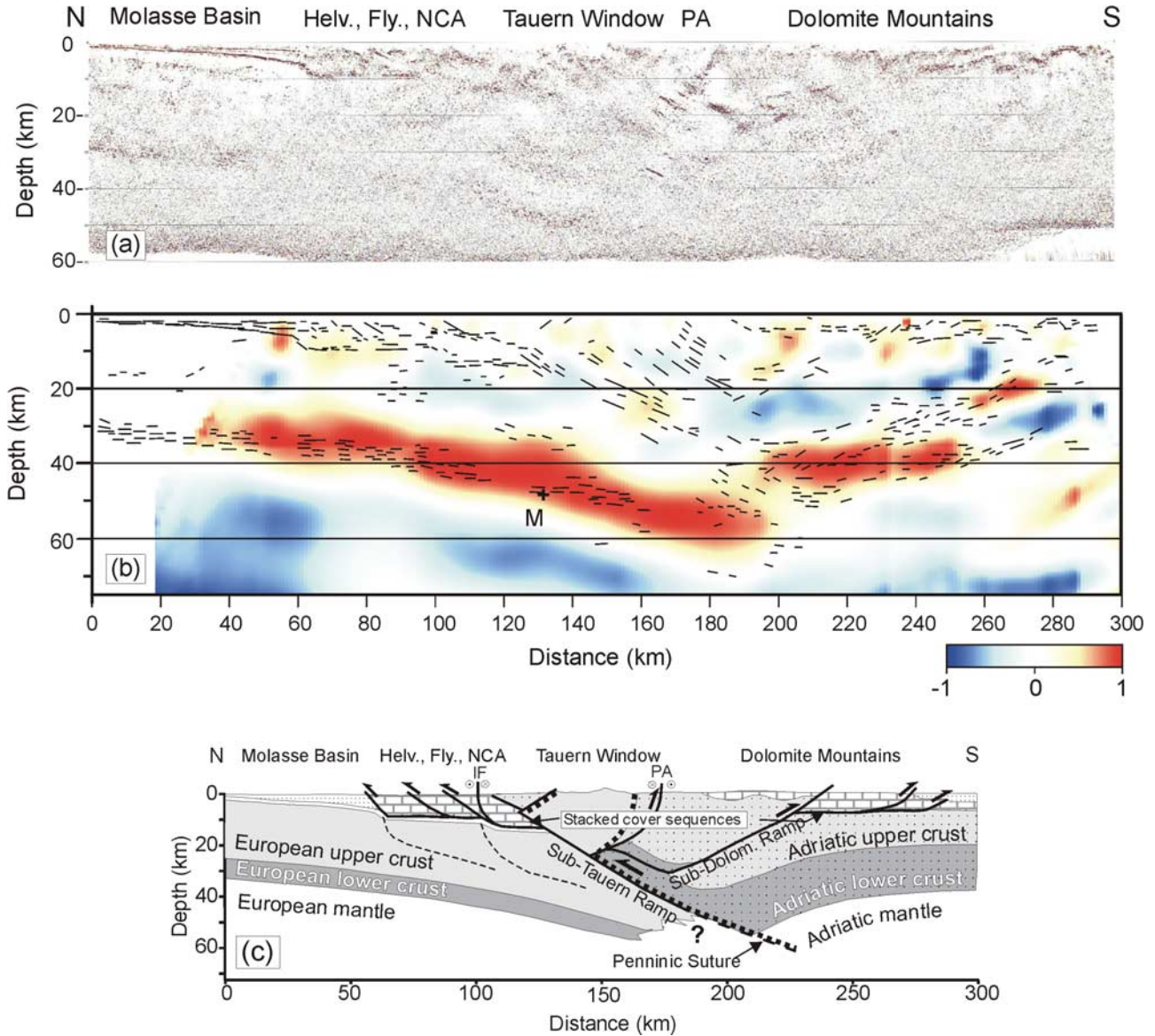
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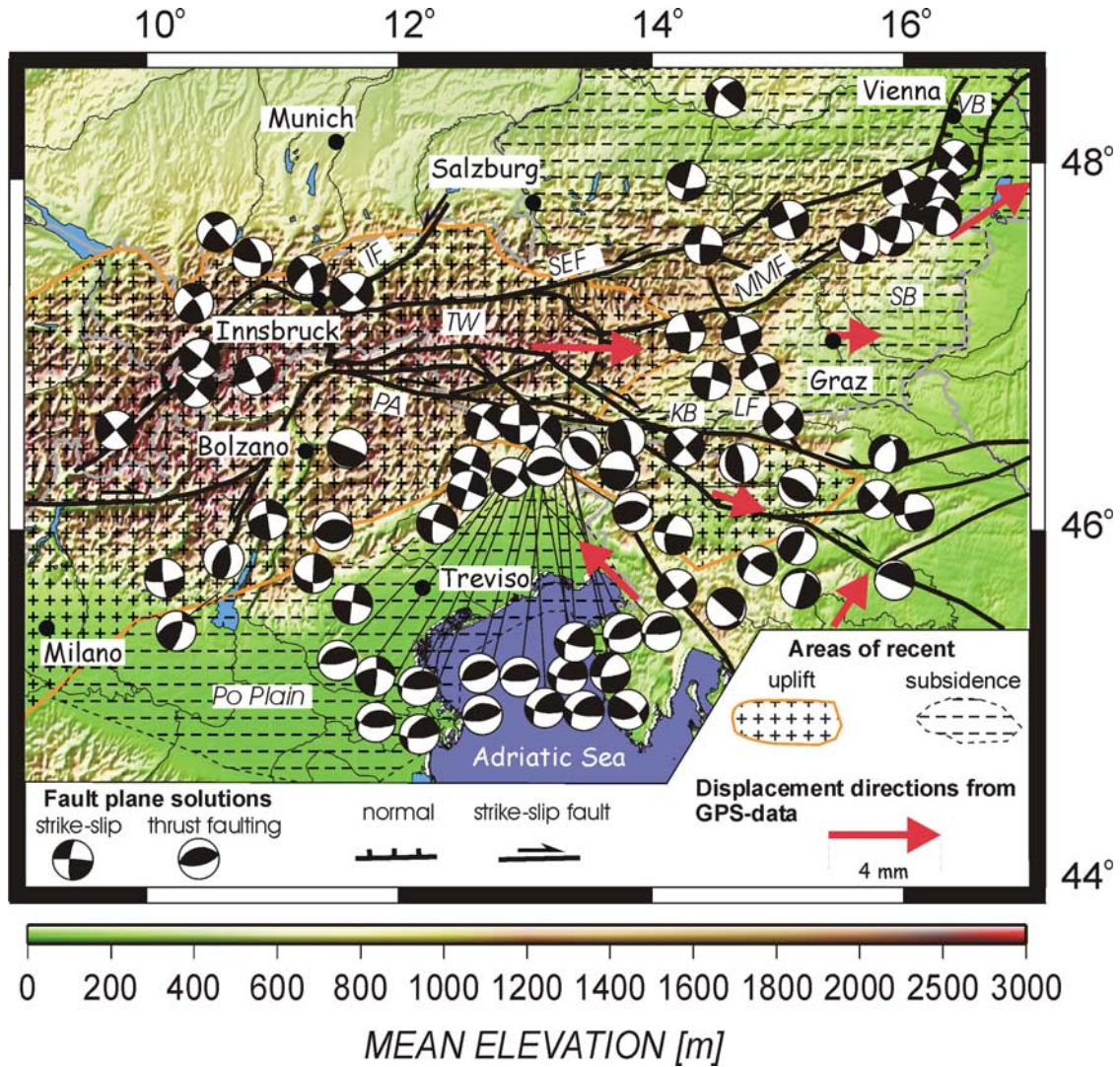
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**Figure 2.** Sections along the TRANSALP transect showing (a) migrated reflection seismic data (vibroseis data); (b) line drawing of reflections deduced from vibroseis and explosive sections superimposed on depth-migrated receiver functions, and (c) simplified interpretation of the seismic data. All figures are taken from *TRANSALP Working Group* [2002]. Note that Figures 2b and 2c have been modified slightly. The large scale structure at the Inntal Fault has been adopted after *Reiter et al.* [2003]. The cross in Figure 2b marks the position of the Moho (M) deduced from refraction seismic data along the ALP75 section [Yan and Mechie, 1989]. Fly, Flysch Zone; Helv., Helvetic Zone; IF, Inntal Fault; NCA, Northern Calcareous Alps; PA, Periadriatic Fault.





**Figure 6.** Summary of neotectonic data of the Eastern Alps. Data are taken from *Arca and Beretta* [1985], *Slejko et al.* [1989], *Kahle et al.* [1997], *Bressan et al.* [1998], *Gerner et al.* [1999], *Reinecker and Lenhardt* [1999], *Grenerczy et al.* [2000], and *Höggerl* [2002]. Note that for the seismically very active regions of northern Italy, Slovenia, and Croatia only a representative selection of focal mechanism can be shown. IF, Inntal Fault; KB, Klagenfurt Basin; LF, Lavanttal Fault; Mur, Mürz Fault; PA, Periadriatic Fault; SB, Styrian Basin; SEF, Salzach-Ennstal Fault; TW, Tauern Window; VB, Vienna Basin.